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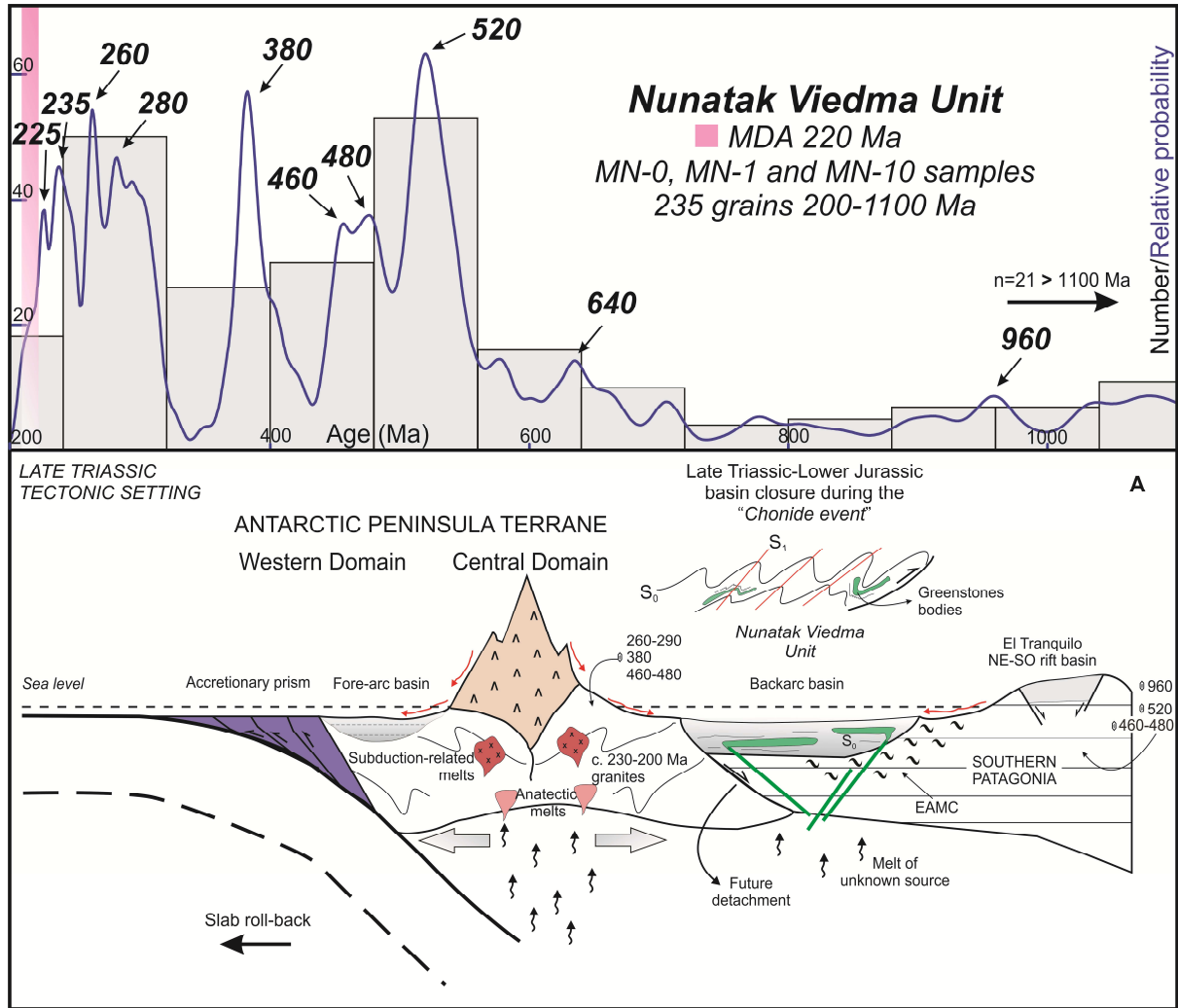
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1 **The metamorphic rocks of the Nunatak Viedma in the Southern**
2 **Patagonian Andes: provenance sources and implications**
3 **for the early Mesozoic Patagonia-Antarctic Peninsula**
4 **connection**

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31 **Abstract.** The Nunatak Viedma within the Southern Patagonian Icefield has been
32 considered as a volcanic center based on its geomorphologic features, despite the fact that
33 field explorations by Eric Shipton determined its metamorphic nature 70 years ago. We
34 carried out fieldwork to characterize this isolated outcrop and performed the first U-Pb
35 dating in detrital zircons from the basement rocks located inside the Southern Patagonian
36 Icefield. We recognized very-low grade metamorphic rocks, corresponding principally to
37 metapelites and metapsammites, and scarce metabasites. Detrital zircons in three
38 metapsammitic samples (composite group of 240 grains) yielded prominent age population
39 peaks at ~1090, ~960, ~630, ~520, ~480-460, ~380, ~290-260, ~235-225 Ma that are
40 typical of Gondwanide affinity, and youngest grains at ~208 Ma. Maximum depositional
41 ages of 225, 223 and 212 Ma were calculated for each sample from the youngest cluster of
42 ages. This distinctive and novelty Late Triassic age justifies differentiate the Nunatak
43 Viedma Unit from the Devonian-early Carboniferous and Permian-Early Triassic (?) belts
44 of the Eastern Andean Metamorphic Complex. Possible primary source areas for the
45 detrital zircons are outcropping in southern Patagonia, the Antarctic Peninsula, and the
46 Malvinas Islands. Additionally, secondary sources could be part of the erosion and
47 recycling of metasediments from the Eastern Andean Metamorphic Complex. We propose
48 that the cluster of Triassic ages is related to the volcanic arc emplaced along the Antarctic
49 Peninsula and active at that timewhen was still attached to southern Patagonia during the
50 Triassic. The dynamics of the early Mesozoic orogen is also discussed.

51 **Keywords:** Nunatak Viedma, very-low grade metamorphic rocks, detrital zircons ages,
52 Antarctic Peninsula, Southern Patagonian Andes, Southwestern Gondwana.

53

54 1. INTRODUCTION

55 The Nunatak Viedma located in the Southern Patagonian Icefield (Fig. 1; 49°22'S,
56 73°19'W), has been considered as an active volcano because of its geomorphologic
57 features, which resemble a group of volcanoes (Lliboutry, 1956; Mazzoni et al., 2010). The
58 early descriptions were based on the first aerial survey that covered the Patagonian icefields
59 (Lliboutry, 1956). Shipton (1960, 1963), however, undertook a field investigation between
60 1958-1962 recognizing that the Nunatak Viedma is entirely composed of metamorphic
61 rocks partially cover by quaternary deposits, mentioning that "*there was no sign whatever*
62 *of any volcanic activity*". Nevertheless, some technical works were recently carried out to
63 characterize the composition and geomorphology of this supposed volcano (Kilian, 1990;
64 Kobashayi et al., 2010). On the other hand, mapping by the Argentine Mining Geological
65 Survey (SEGEMAR; Giacosa et al., 2012a) assigned this out crop to the late Paleozoic
66 Bahía de la Lancha Formation. By last, Blampied et al. (2012) described it as a folded
67 metasedimentary sequence composed of schists and gneisses, assigning their particular
68 geomorphological volcanic-like shape to glacial processes.

69 Metamorphic rocks of the Southern Patagonian Andes principally include low-grade,
70 metasedimentary units grouped in two distinctive, north-south oriented, Eastern and
71 Western complexes flanking the Patagonian Batholith (Hervé et al., 2003; Hervé et al.,
72 2008; Calderón et al., 2016). Deposition of the protoliths and subsequent metamorphism
73 and deformation occurred between the late Paleozoic and the early Mesozoic (Hervé et al.,
74 2008). In the widespread distributed Eastern Andean Metamorphic Complex (EAMC), the
75 protoliths were deposited during the Late Devonian-early Carboniferous and Late Permian-
76 Early Triassic (Hervé et al., 2003; Augustsson et al., 2006) and metamorphosed, at least the
77 older units, during the Gondwanide orogeny (Thomson and Hervé, 2002; Giacosa et al.,

78 2012; Heredia et al., 2016). In this vast metamorphic complex were included the Bahía de
79 la Lancha and Río Lácteo formations defined in the Argentinean territory (Hervé et al.,
80 2008). A regionally widespread angular unconformity with the overlying Upper Jurassic
81 volcano-sedimentary rocks along the Patagonian hinterland (Giacosa and Heredia, 2004;
82 Ghiglione et al., 2009; Giacosa et al., 2012; Zerfass et al., 2017), gives a minimum age for
83 deformation of the folded basement units.

84 On the other hand, units of the western Metamorphic Complexes exhibit a geological
85 history related to subduction dynamics and accretion of exotic terranes during the Triassic-
86 Jurassic (Hervé et al., 2003; Hervé et al., 2008; Hervé and Fanning, 2001; Willner et al.,
87 2009; Angiboust et al., 2017).

88 From correlation of Paleozoic-early Mesozoic magmatic and tecto-metamorphic
89 events (Pankhurst et al., 2003; Hervé et al., 2006; Castillo et al., 2016; Heredia et al., 2016;
90 Riley et al., 2016; González et al., 2018), it has been proposed that South America,
91 Patagonia, and the Antarctic Peninsula were contiguous regions forming the SW margin of
92 Gondwana. In this sense, several authors have discussed the spatial relation between
93 Patagonia and the Antarctic Peninsula (Fig. 2; Suárez, 1976; Harrison et al., 1979; Lawver
94 et al., 1998; Hervé and Fanning, 2003; Hervé et al., 2006; Ghidella et al., 2007; Calderón et
95 al., 2016; Heredia et al., 2016) conforming the Terra Australis Orogen before Gondwana
96 breakup (Fig. 2; Suárez, 1976; Harrison et al., 1979; Miller et al., 1987, 2007; Lawver et
97 al., 1992, 1998; Jokat et al., 2003; König and Jokat, 2006; Ghidella et al., 2007). The "tight-
98 fit" model of Lawver et al. (1998) suggests that the northern edge of the Antarctic
99 Peninsula was located near the current Golfo de Penas (Fig. 2), although clear correlations
100 between Patagonian and the Antarctic Peninsula are still missing to validate this hypothesis

101 (Hervé et al., 2006). After the onset of oceanic spreading, lateral displacements and block
102 rotations during Mesozoic-Cenozoic times, including opening of the Scotia plate,
103 determined the current plate configuration and the distribution of the continental fragments
104 of Gondwana (Suárez, 1976; Harrison et al., 1979; Jokat et al., 2003; Eagles, 2016;
105 Ghiglione et al., 2016).

106 In this contribution, we provide a lithological characterization and U-Pb zircon
107 detrital ages from metasedimentary rocks of the Nunatak Viedma and discuss its evolution
108 within the geodynamic context of the SW edge of Gondwana. The distinctive Late Triassic
109 age justifies to differentiate a Nunatak Viedma Unit (NVU) separate from the EAMC.
110 Furthermore, we discuss the possible sedimentary sources for the protolith of the NVU and
111 its implications for the paleogeography and location of the Antarctic Peninsula with respect
112 to Patagonia during Triassic times.

113 **2. GEODYNAMIC CONTEXT**

114 The Terra Australis Orogen developed along the active margin of Gondwana during
115 Neoproterozoic-early Mesozoic times (Fig. 2; Cawood, 2005; Cawood and Buchan, 2007).
116 At the end of the Paleozoic, several events of terrane accretion are recorded in South
117 America (Pankhurst et al., 2006; Ramos, 2008; Calderón et al., 2016), between these the
118 accretion of the Antarctic Peninsula terrane against Patagonia (Ramos, 2008; Calderón et
119 al. 2016; Heredia et al. 2016), which indicate the final stages of Pangea assembly (320-250
120 Ma; Cawood and Buchan, 2007). Early Mesozoic tecto-magmatic events associated with
121 continental breakup, are recorded along its SW margin (Bruhn et al., 1978; Pankhurst et al.,

122 2000; Riley et al., 2001; Calderón et al., 2016; González et al., 2016), related to extensional
123 movements that started to separate the Antarctic Peninsula from southern Patagonia.

124 Early Mesozoic paleogeographic reconstructions suggest that the Antarctic Peninsula
125 was located in a very close position, or even attached, to SW Patagonia (Harrison et al.,
126 1979; Miller et al., 1987; Lawver et al., 1992, 1998; Jokat et al., 2003; Hervé et al., 2006;
127 König and Jokat, 2006; Ghidella et al., 2007; Heredia et al., 2016). However, these models
128 still differ in the exact paleo-latitudinal position of the Antarctic Peninsula along the Pacific
129 margin during the Triassic or older times. Some hypothesis suggests that the northern
130 margin of Antarctic Peninsula was located nearby Golfo de Penas (Fig. 2; Lawver et al.,
131 1998; Hervé and Fanning, 2003; König and Jokat, 2006) or farther north, close to the 39°S,
132 according to Heredia et al. (2016), while other reconstructions propose that the peninsula
133 was situated either east of Patagonia or in continuity towards the south (Suárez, 1976;
134 Miller, 2007).

135 Plate reconstructions from aeromagnetic data establish a southward drifting of the
136 Antarctic Peninsula during the Jurassic (Jokat et al., 2003; König and Jokat, 2006),
137 synchronous with rifting and oceanic floor spreading in the Rocas Verdes basin in the
138 backarc region of Patagonia (Dalziel, 1981; Calderón et al., 2007a).

139 **2.1. Late Paleozoic-early Mesozoic metamorphic units from southernmost** 140 **South America and Antarctic Peninsula.**

141 These units include several lower Paleozoic-Mesozoic blocks that formed the SW
142 edge of Gondwana including possibly exotic terranes (Thomson and Hervé, 2002; Millar et
143 al., 2002; Hervé et al., 2003, 2007, 2008; Calderón et al., 2016).

144 **2.1.1. Southern Patagonian Andes**

145 Metamorphic rocks cropping out in the Southern Patagonian Andes (Fig. 3) can be
146 divided into (1) the **EAMC (Eastern Andes Metamorphic Complex)** along the eastern
147 flank of the Patagonian Batholith, (2) the **coastal accretionary complexes in the Pacific**
148 **archipelago**, and to a lesser extent (3) the **ophiolitic complexes of the Rocas Verdes**
149 **basin**.

150 (1) The **EAMC** crops out along the inner core of the South Patagonian Andes
151 (Hervé et al., 2008), including several units that compose the thick-skinned basement
152 domain towards the east (Ghiglione et al., 2009). The Bahía de la Lancha and Río Lácteo
153 formations, defined in Argentinean territory (see Giacosa and Márquez, 2002) were
154 included by Hervé et al. (2008) in this vast metamorphic unit. The EAMC mainly consists
155 in turbiditic sequences with occasional limestone bodies and metabasalt and their
156 metamorphic equivalents in greenschist facies (Hervé et al., 2008). Sedimentary protoliths
157 were deposited either in a passive margin environment (Augustsson and Bahlburg, 2003a,
158 2003b, 2008; Lacassie, 2003) or in a forearc setting (Permuy Vidal et al., 2014; Calderón et
159 al., 2016) during the Late Devonian-early Carboniferous (Augustsson et al., 2006) and
160 metamorphosed previous to the Permian (Thomson and Hervé, 2002). Giacosa et al.
161 (2012b), propose that the Bahía de la Lancha Formation was deformed during the late
162 Paleozoic Gondwanide orogeny. Additionally, the western belt of the EAMC exhibits Late
163 Permian-Early Triassic ages (Hervé et al., 2003; Augustsson et al., 2006), however its
164 deformation and metamorphism age is still unknown. High-grade metamorphic rocks and
165 migmatites located on the western border of the EAMC resulted from the emplacement of

166 Late Jurassic granitoids of the South Patagonian Batholith (Hervé et al., 2007; Calderón et
167 al., 2007b).

168 **(2) The Coastal accretionary-subduction complexes** (Fig. 3) include (see Hervé *et*
169 *al.*, 2008, Calderón et al., 2016) the CMC (Chonos Metamorphic Complex), the MDAC
170 (Madre de Dios Accretionary Complex) and the DAMC (Diego de Almagro Metamorphic
171 Complex).

172 The CMC from the Chonos Archipelago is composed of metaturbidites and
173 metabasites and meta-cherts in minor quantities (Hervé et al., 2008). The protolith was
174 deposited during the Late Triassic (Fang et al., 1998; Hervé and Fanning, 2001) and
175 subsequently affected by metamorphism and deformation during Late Triassic-Early
176 Jurassic times, possibly related to the Chonide deformational event (Hervé et al., 2006,
177 2008). The high-pressure/low-temperature metamorphism is typical of accretionary prisms
178 (Willner et al., 2000; Ramírez-Sánchez et al., 2005).

179 The MDAC includes three lithostratigraphic units known as: (i) the Tarlton
180 limestone, composed by a massive pelagic limestone body deposited during the late
181 Carboniferous-early Permian, (ii) the Denaro Complex represented by pillow basalts and
182 radiolarian cherts, and (iii) the Duque de York Complex composed by turbiditic
183 successions unconformably deposited over the Tarlton limestone and Denaro Complex
184 (Forsythe and Mpodozis, 1979, 1983). The lower-middle Permian Duque de York Complex
185 (Faundéz et al., 2002) experienced low-grade metamorphic conditions during the Early
186 Jurassic, possibly related to an accretionary event (Thomson and Hervé, 2002). The Denaro

187 and Tarlton complexes have generally been interpreted as oceanic exotic terranes (Faundéz
188 et al., 2002; Rapalini et al., 2001).

189 The DAMC corresponds to the assemblage of two metamorphic units tectonically
190 juxtaposed by the Puerto shear zone: the Almagro HP-LT Complex composed by
191 metamafic volcanic rocks with scarce metasedimentary rocks and the Lazaro Unit with
192 seafloor-derived metamafic and metasedimentary rocks (Angiboust et al., 2017). The
193 protolith of these high-pressure rocks have been deposited in the Late Jurassic and
194 metamorphosed as part of an accretionary wedge that was active during the Cretaceous
195 (Hervé and Fanning, 2003; Hyppolito et al., 2016; Angiboust et al., 2017, 2018).

196 **(3) Ophiolitic complexes of the Rocas Verdes basin** consist of sequences of
197 massive and layered gabbros, sheeted dykes and basalts with oceanic affinity, whereas
198 ultramafic rocks are absent (Stern and De Witt, 2003; Calderón et al., 2007a). From north
199 to south they are the Sarmiento Complex in the Southern Patagonian Andes and the Capitán
200 Aracena and Tortuga complexes in the Fuegian range (Stern and De Witt, 2003; Calderón
201 et al., 2013). These rocks were emplaced during the Late Jurassic-Early Cretaceous in a
202 backarc rift basin (Calderón et al., 2007). The subsequent basin closure and obduction
203 occurred during the Cenomanian-Coniacian (Calderón et al., 2007a; Klepeis et al., 2010;
204 Calderón et al., 2012), producing a major advance of the orogenic front and a wedge-top
205 setting for the adjacent Austral-Magallanes basin depocenter (Fosdick et al., 2011;
206 Likerman et al., 2013; Ghiglione et al., 2014, 2016).

207 **2.1.2. Tierra del Fuego**

208 Metamorphic rocks of the Cordillera Darwin and extra-Andean region are known as
209 **Cordillera Darwin Metamorphic Complex** and **Tierra del Fuego Igneous and**
210 **Metamorphic Complex** (Fig. 3), respectively. At Cordillera Darwin the clastic protolith of
211 the metasedimentary units were deposited in the Paleozoic (Hervé et al., 1981a, 2010b;
212 Barbeau et al., 2009b) and their detrital zircon age spectra are comparable with those of the
213 EAMC (Barbeau et al., 2009b; Hervé et al., 2010b). These metamorphic rocks are
214 unconformably overlain by Jurassic volcanoclastic rocks, and the whole succession has been
215 strongly deformed and metamorphosed in the Cretaceous by the Andean Orogeny (Hervé et
216 al., 2010b; Maloney et al., 2011). The extra-Andean region of Tierra del Fuego is
217 characterized by Cretaceous–Cenozoic sedimentary rocks of the Austral-Magallanes basin
218 (e.g. Ghiglione et al., 2010). Underlying foliated plutonic rocks (orthogneisses) and
219 migmatitic gneisses were detected in boreholes. These rocks have Cambrian crystallization
220 ages and are affected by a late Permian magmatic-anatectic event (Sölner et al., 2000;
221 Calderón et al., 2010; Hervé et al., 2010a; Castillo et al., 2017).

222 **2.1.3. Antarctic Peninsula**

223 Basement rocks from the Antarctic Peninsula at Graham Land and Palmer Land
224 yield *in situ* ages ranging from Ordovician to Late Triassic (Fig. 3; Millar et al., 2002; Riley
225 et al., 2012).

226 In this region, three geological domains separated by shear zones are recognized
227 (Fig. 3; Vaughan and Storey, 2000; Ferraccioli et al., 2006; Burton-Johnson and Riley,
228 2015). The Permian-Upper Triassic to Lower Jurassic Central domain is composed by
229 subduction-related granites, orthogneiss and migmatites cropping out in Palmer Land area

230 (Wever et al., 1994; Millar et al., 2002; Riley et al., 2012). The Eastern domain includes the
231 Permian to Triassic **TPG (Trinity Peninsula Group)**, whose age is constrained by U-Pb
232 dating in detrital zircons and stratigraphic relationships (Carvalho et al., 2005; Barbeau et
233 al., 2009a; Castillo et al., 2016). This group has been divided into the Hope Bay, View
234 Point and Legoupil formations composed of metaturbidites and metapillow lavas deposited
235 in a forearc basin (Hyden and Tanner, 1981). Unconformably overlain on the TPG, Heredia
236 et al. (2016) redefine the uppermost Düse Bay Formation described by Del Valle et al.
237 (2007) as a synorogenic succession deposited in a retroarc foreland basin and related to the
238 Tabarin Permian-Triassic orogeny. To the northeast of the Antarctic Peninsula, the South
239 Orkney Island (Fig. 3) at the southern limb of the Scotia Range exhibits outcrops of the
240 **Greywacke-shale Formation** interpreted as a possible Triassic equivalent of the TPG
241 (Dalziel, 1982; Trouw et al., 1997).

242 **3. METAMORPHIC ROCKS OF THE NUNATAK VIEDMA UNIT**

243 **3.1. Field and petrographic lithological characterization**

244 The rocky out crop of the Nunatak Viedma is modeled by glaciary processes and
245 presents no evidence of volcanic activity. The rocks correspond mainly to metapsammites,
246 metapelites and scattered interleaved metabasites (Fig. 6). The metapsammites are gray,
247 present scarce clasts of both felsic volcanic rocks and granites and pumice wisps with tube
248 vesicle texture. In all samples, grains of quartz, white mica, and plagioclase are dominant
249 (Fig. 7A) with scarce K-feldspar and accessories like zircon, tourmaline, apatite, and
250 titanite. The minerals are distributed in the quartz-plagioclase and quartz-white mica
251 domains, which are interpreted as remnants of a prior sedimentary lamination (S₀).

252 Sometimes the mechanically reoriented and kinked detrital mica together with dissolution
253 films define a diagenetic foliation (*sensu* Passchier and Trouw, 2005). The metamorphic
254 mineral assemblage is composed of quartz, chlorite, white mica and carbonate (Fig. 7B).
255 The fine bands formed by intergrowths of white mica and chlorite define the S_1 foliation
256 (Fig. 7B, 7C). Chlorite locally replaces white mica and also is present as porphyroblasts.
257 The mica-rich domains, consisting of white mica and chlorite, are affected by a poorly
258 developed crenulation cleavage (S_2 ; Fig. 7C). By last, the rocks are crosscut by veinlets of
259 carbonate and quartz.

260 The metabasites preserve amygdaloidal and porphyritic primary textures (Fig. 7D).
261 Primary igneous phases are composed of plagioclase, clinopyroxene, pseudomorphs of
262 olivine and opaques (Fig. 7D). Metamorphic mineral assemblages vary between samples;
263 the most common are: i) chlorite, smectite, zeolite and carbonates (Fig. 7D); ii) chlorite,
264 pumpellyite (?); iii) actinolite, chlorite; and iv) smectite, quartz, and carbonate. These
265 minerals occur in amygdules and micro-veinlets, or as the alteration product of plagioclase.

266 **3.2. Estimations of metamorphic P-T conditions**

267 Deformation processes associated to micro-structures, identification of metamorphic
268 mineral assemblages and their comparison with petrogenetic grids for metabasites
269 (Schiffman and Day, 1999) and with typical metamorphic mineral assemblage for
270 metapelites (Bucher and Grapes, 2011), indicate that metamorphism during deformation
271 and developments of S_2 foliation progressed in very low to low-grade conditions.

272 The textural equilibriums among white mica + chlorite + quartz + carbonate in
273 metapsammites indicate a sub-greenschist facies of metamorphism (chlorite zone), with
274 estimated temperatures at 200-300° C and pressures between 1.5 and 3.0 kbar.

275 The mineral assemblage identified in the metabasites indicate a formation
276 temperature of 180-230° C and variable pressures between 1 and 3 kbar to chlorite +
277 pumpellyite mineral association, while the appearance of actinolite could be point out
278 temperatures greater than 300° C. These physical conditions correspond to the transition
279 between sub-greenschist and greenschist facies.

280 **4. U-PB ZIRCON DETRITAL AGES IN METASEDIMENTARY ROCKS**

281 **4.1. Analytical techniques**

282 Three samples were taken in site 49°25'50"S/ 73°18'30"W (Fig. 1) to analyze
283 detrital zircons by U-Pb method in the Washington State University Laboratory, following
284 the methodology described in Chang et al. (2006). Approximately three hundred zircon
285 grains were separated from a total of 5 kg of three metapsammites (samples MNO, MN1
286 and MN10). Heavy mineral fractions were concentrated and separated into 100, 150 and
287 250 mm size fractions by standard crushing and panning. Zircon fractions of roughly 400
288 grains were handpicked in alcohol under a binocular microscope for geochronology
289 analysis. Zircons of unknown ages and standards were handpicked under the microscope
290 and mounted in a 1-inch diameter epoxy puck and slightly ground and polished to expose
291 the surface and keep as much material as possible for laser ablation analyses. After
292 cathodoluminescence imaging, the LA-ICP-MS U-Pb analyses were conducted using a
293 New Wave Nd: YAG UV 213-nm laser coupled to a Thermo Finnigan Element 2 single

294 collector, double-focusing, magnetic sector ICP-MS. Laser spot size and repetition rate
295 were 30 microns and 10 Hz, respectively. He and Ar carrier gases delivered the sample
296 aerosol to the plasma. Each analysis consists of a short blank analysis followed by 250
297 sweeps through masses 202, 204, 206, 207, 208, 232, 235, and 238, taking approximately
298 30 s. Time-independent fractionation was corrected by normalizing U/Pb and Pb/Pb ratios
299 of the unknowns to the zircon standards (Chang et al., 2006). U and Th concentration were
300 monitored by comparing to NIST 610 trace element glass. Two zircon standards were used:
301 Plesovice, with an age of 338 Ma (Sláma et al., 2008) and FC-1, with an age of 1099 Ma
302 (Paces and Miller, 1993). Uranium-lead ages and plots were calculated using Isoplot
303 (Ludwig, 2003). Analyses were corrected assuming concordance and applying a common
304 Pb correction using the 207 Pb method (Williams, 1998).

305 The analytical data with concordant ages are reported in the **Supplementary Data**,
306 including uncertainties at the 1σ level, and measurement errors. Best ages were chosen
307 based on the precision of the isotopic systems, thus $^{206}\text{Pb}/^{238}\text{U}$ ages were selected for
308 zircons younger than 1200 Ma and $^{207}\text{Pb}/^{206}\text{Pb}$ ages for older ones (Gehrels et al., 2008). In
309 a first step, all ages were plotted in the Tera and Wasserburg (1972) Concordia diagram
310 (Fig. 4). Afterward, a filtering was applied to assess discordant ages, which can be
311 quantified using the ratio between $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ages (Fig. 4; Wetherill, 1956;
312 Spencer et al., 2016). The resulting percentage of “discordance” was set at 20% as
313 commonly used for detrital zircons (e.g. Nelson and Gehrels, 2007; Naipauer et al., 2010;
314 among many others).

315 4.2. Estimation of maximum depositional age

316 Obtaining the MDA (maximum depositional age) of a geological unit is one of the
317 most widely used applications of detrital zircon geochronology and can be calculated from
318 different methods (Dickinson and Gehrels, 2009). The age of the youngest zircon from a
319 data set gives some insights, but uncertainty on the precision of a single age is still large
320 since it does not allow a statistical or modal test (Dickinson and Gehrels, 2009; Gehrels,
321 2014). Other methodologies provide more conservative estimations and have turned out
322 coherent in units with biostratigraphic control, as the weighted mean age of a cluster
323 composed by the three (or more) youngest zircons whose σ errors overlap (Dickinson and
324 Gehrels, 2009). We applied this methodology to calculate a conservative MDA from each
325 sample, and the average of these three ages was assumed as the MDA of the unit.

326 **4.3. Detrital zircon features and U-Pb data**

327 In all the analyzed samples the pattern of detrital ages, external morphology of the
328 detrital zircon grains and Th/U ratio are very similar. For this reason, they will be described
329 as a unique set. Figure 5 shows the distribution of detrital zircon ages for each sample and
330 the probability plots and associated maximum depositional age estimations.

331 About one hundred zircon grains were analyzed from each sample but only 79 of 101
332 (MN0), 94 of 112 (MN1) and 83 of 100 (MN10) zircons grains with discordance values
333 <20% endured filtering. Concordant zircon ages define a complex pattern of detrital ages
334 that range from the Archean to the Late Triassic (Fig. 5).

335 Detrital zircon grains are 50-260 μm in length (median length is 114 μm) and their
336 external morphology exhibits different classes (according to Gärtner et al., 2013),
337 concentrated in stubby- and stalky-types. The Th/U ratios show values between 0.05-2.47;

338 in all samples the Th/U is at least 87% greater than 0.2, indicating an igneous origin for
339 most of them.

340 The age distribution of detrital zircon grains include representative Paleozoic-early
341 Mesozoic (65%), Proterozoic (34%) and isolated Archean ages (1%; Fig. 5). The peaks of
342 detrital ages define five main groups in the Paleozoic-Mesozoic: Early Cambrian (~520
343 Ma), Early-Middle Ordovician (~480-460 Ma), Late Devonian (~380 Ma), Permian (~290-
344 260 Ma) and Triassic (~235-225 Ma). Precambrian ages are concentrated in the
345 Neoproterozoic around ~1090 Ma, ~960 Ma, and ~630 Ma. MDAs are Late Triassic, at 225
346 Ma for MN-0, 212 Ma for MN-1 and 223 Ma for MN-10 (Fig. 5), with errors within ± 5
347 Ma.

348 5. DISCUSSION

349 In a first approach, the metamorphic rocks of the Nunatak Viedma could be
350 included as part of the EAMC on the basis of its geographical proximity and lithological
351 similarities. However, the Late Triassic peak of U-Pb zircon detrital ages clearly
352 differentiates a new tectonostratigraphic unit in the *collage* of pre-Andean metamorphic
353 belts of the Southern Patagonian Andes. We propose to denominate these rocks as Nunatak
354 Viedma Unit (NVU). Based on petrographic and geochronological data, the geological
355 significance of the NVU and its possible correlations with other metamorphic complexes of
356 the southwestern edge of Gondwana in the early Mesozoic, are discussed below.

357 5.1. Provenance sources and paleogeographic implications

358 In quartz-rich metapsammites, with a fine-grained terrigenous, probably turbiditic
359 protolith, detrital zircon populations of Late Triassic, Paleozoic (Cambrian, Ordovician,
360 Devonian, Permian), Neoproterozoic and isolated Archean ages were identified.

361 Archean and Proterozoic detrital zircon ages have been previously identified in
362 metamorphic complexes located along the Patagonian Andes, for example in the EAMC
363 (Hervé et al., 2003; Augustsson et al., 2006). However, Patagonia as a primary (igneous)
364 cratonic source of detrital zircons still remains under discussion, because of the lack of
365 magmatic outcrops of that age. Precambrian primary source areas are located in South
366 America north of Patagonia, Africa and East Antarctica (cf. Hervé et al., 2003; Augustsson
367 et al., 2006), and we consider these primary sources areas to be located too far north and
368 east to have reach the NVU basin directly. For this reason, Precambrian detrital ages
369 identified within the Triassic NVU, can be more easily explained by recycling of the nearby
370 EAMC and also from other Paleozoic igneous-metamorphic complexes with outcrops in
371 surrounding areas, probably exposed at the time of NVU deposition.

372 Recent works consider the Malvinas Islands and southern Patagonia as a single
373 continental block (Ramos et al., 2017; Schilling et al., 2017). From that perspective,
374 Mesoproterozoic Grenvillian sources could correspond to granites and gneisses of the Cabo
375 Belgrano Complex, which are cropping out in the Malvinas islands (Fig. 3b; Cingolani and
376 Varela, 1976; Rex and Tanner, 1982; Thomas et al., 2000) and possibly in the neighboring
377 Argentinean Atlantic platform (Wareham et al., 1998; Chemale Jr. et al., 2018).

378 On the other hand, Paleozoic and Mesozoic detrital zircons could potentially come
379 from multiple sources (Fig. 3a, b). The early Cambrian zircon ages, related to the Pampean

380 orogen in center and northwest Argentina and the Ross-Delamerian orogen in northern
381 Patagonia (Ramos and Naipauer, 2014; González et al., 2018) and Antarctic-Australia
382 (Cawood and Buchan, 2007). In the North Patagonian Massif, the low-grade (Nahuel Niyeu
383 and El Jagüelito formations) and high-grade (Mina Gonzalito Complex) metamorphic units
384 and the Tardugno Granodiorite present Cambrian ages (Fig. 3b; Pankhurst et al., 2006;
385 González et al., 2011; Rapalini et al., 2013; Greco et al., 2017). Cambrian gneisses and
386 foliated plutonites, pertaining to the Tierra del Fuego Igneous and Metamorphic Complex,
387 have been identified in exploratory boreholes of the Austral-Magallanes basin (Söllner et
388 al., 2000; Hervé et al., 2010a). The crystallization age of these igneous rocks was calculated
389 at around of 520 Ma (Söllner et al., 2000; Pankhurst et al., 2003; Hervé et al., 2010a), it
390 turns out to be highly coincident with early Cambrian peak presents in our samples.

391 Detrital ages grouped at ~460-480 Ma peaks could be sourced from Lower to Middle
392 Ordovician granitoids of the Punta Sierra Plutonic Complex in the North Patagonian Massif
393 (Fig. 3b; Pankhurst et al., 2006; González et al., 2008, 2014) and the Río Deseado Complex
394 in the Deseado Massif (Loske et al., 1999; Pankhurst et al., 2003). Also, basement outcrops
395 in Graham Land at the Antarctic Peninsula are composed of dioritic gneisses with
396 Ordovician ages (Fig. 3a; Riley et al., 2012).

397 The Devonian peak could be associated with widely represented igneous rocks in the
398 Northern Patagonian Andes (Varela et al., 2005; Pankhurst et al., 2006; Hervé et al., 2016)
399 and in the western part of the North Patagonian Massif (Hervé et al., 2016). These ages are
400 poorly represented in the Southern Andes, Antarctic Peninsula (Millar et al., 2002;
401 Pankhurst et al., 2003) and in the Deseado Massif (Pankhurst et al., 2003). Devonian

402 detrital ages are also common in the EAMC (Augustsson et al., 2006) and in metamorphic
403 rocks from the western Deseado Massif (Permuy Vidal et al., 2014).

404 An important Permian-Triassic zircon population suggests erosion of post-
405 Gondwanan igneous rocks and of a late Paleozoic-early Mesozoic magmatic arc active
406 during sedimentation of the NVU. An alternative explanation to the Permian detrital ages
407 could be the recycling from the western belt of the EAMC. Permian Igneous rocks (~280-
408 250 Ma) has been identified in the North Patagonian Massif (Fig. 3b; Pankhurst et al.,
409 2014), although plutonic samples of that age are scarce in southern Patagonia. Recently,
410 new zircon U-Pb radiometric data obtained principally in the Antarctic Peninsula, but also
411 some in Tierra del Fuego (Calderón et al., 2010; Hervé et al., 2010a; Millar et al., 2012;
412 Castillo et al., 2017) demonstrates a widespread distributed Permian metamorphic-
413 magmatic event. Castillo et al. (2016) studied Permian to Triassic metasedimentary rocks
414 from the Pacific coast of SW Chile using isotope analyses in detrital zircon grains and they
415 suggested the Permian subduction-related magmatic arc located in Patagonia and west
416 Antarctic as an important source.

417 From a paleogeographic point of view, plutonic rocks (between 236-200 Ma by
418 Millar et al. 2002 and Riley et al. 2012) from the Antarctic Peninsula arc, located to the W
419 and SW (Fig. 3a) were the closest possible source for Triassic detrital zircons (235-208 Ma)
420 during sedimentation of the protolith of the NVU, taking into account the hypothesis that
421 the Antarctic Peninsula was attached to Patagonia at that time (Fig. 2; Calderón et al. 2016;
422 Castillo et al., 2016; Heredia et al., 2016).

423 In summary, U-Pb analysis in detrital zircons from the NVU show a wide distribution
424 of ages with all Paleozoic to early Mesozoic periods very well represented, as well as an
425 important Proterozoic component (Figs. 5 and 8). The diverse although equitable zircons
426 distribution (Fig. 5) can only be explained by considering a multiplicity of sources for the
427 Proterozoic-Paleozoic components, located in Patagonia and Antarctic Peninsula (Fig. 3a,
428 b). We propose that the very well represented and particular Permian-Triassic group of ages
429 (Figs. 5 and 8) was sourced from the active magmatic arc emplaced onto the Trinity
430 Peninsula Group along Antarctic Peninsula.

431 **5.2. Nature, age, and correlations of the Viedma basin**

432 If we consider the petrography, and particularly the detrital zircon age spectra
433 (*sensu* Cawood et al., 2012) as an expression of the tectonic setting of the NVU protolith,
434 i.e. the Viedma basin, their distribution indicate a basin flanked by an active Late Triassic
435 magmatic arc (Fig. 9). In order to further define the nature of this basin and its position
436 with respect to the recognized Triassic magmatic arcs is necessary to take into account the
437 regional correlation with other contemporaneous metamorphic units.

438 Metamorphic belts with protoliths of Triassic age are rare in the Patagonian Andes,
439 except by the CMC located in the present forearc about 300-400 km to the northwest of the
440 NVU (Fig. 3). Detrital zircons ages (Hervé and Fanning, 2001) and fossils record (Fang et
441 al., 1998), constraint the sedimentation age in a Late Triassic basin, within an environment
442 of accretionary wedge (Hervé et al., 1981b; Hervé, 1988), here onwards called the “Chonos
443 basin”.

444 The distribution pattern of detrital ages of the CMC and NVU shows that both
445 basins were mainly fed by Permo-Triassic sources, although the NVU exhibits particular
446 Neoproterozoic, Cambrian, and Ordovician detrital ages probably sourced from the
447 Patagonian massifs (see section 5.1.). Trench or forearc basins such as the Chonos basin,
448 present a large population of detrital ages close to the sedimentation age without large
449 proportions of older zircon grains, while basins in a retroarc position present an input of
450 older detrital sources from the adjacent basement outcrops (Fig. 9; Cawood et al., 2012).
451 Given the broad range of detrital ages found in the NVU, it seems that the Viedma basin
452 was probably a retroarc position (Fig. 10).

453 Conscious of the difficulty that represent to infer paleogeography from provenance
454 data due to the complexity of the drainage systems and their catchment areas (Cawood et
455 al., 2003), the Triassic volcanic arc emplaced along Antarctic Peninsula, and possibly with
456 continuation in Central Patagonia (i.e. Central Patagonian Batholith and Curaco Plutonic-
457 Volcanic Complex, with ages between 222-206 Ma; Rapela et al., 1992; Saini Eidukat et
458 al., 2004; Zaffarana et al., 2014), could have played the role of topographic barrier between
459 both compared basins. In this way, the Viedma basin caught Neoproterozoic, Cambrian and
460 Ordovician zircon grains sourced from extra-andean Patagonian basement highs sources
461 located to the east, while Chonos basin was fed by the flanking Triassic arc and the
462 underneath host rock.

463 **5.3. Early Mesozoic dynamics of the subduction system**

464 Accretionary orogens were traditionally classified according with their surrounding
465 subduction style, considering and linking mainly slab dip angle, strain class in the upper

466 plate and trench migration (Uyeda and Kanamori, 1979; Jarrard, 1986; Stern, 2002; Heuret
467 et al., 2007; Ramos, 2010). In this context, two end-members have been identified:
468 advancing and retreating orogens (*sensu* Cawood et al., 2009). In the second case, the
469 dynamics of the orogen (see more in Collins, 2002a) induce lithospheric extension and the
470 development of backarc basins (Collins, 2002b; Cawood et al., 2009). The early Mesozoic
471 arc-backarc system established along of the west margin of Patagonia and Antarctic
472 Peninsula shows close relationship with this kind of accretionary orogen undergoing
473 extension.

474 Early Mesozoic orthogneisses and foliated granites emplaced along Antarctic
475 Peninsula (mainly in the Central Domain) were interpreted as products derived from the
476 subduction zone but during a regime of lithospheric extension (Wever et al., 1994).
477 Likewise, the Late Triassic in Patagonia and along of the western margin from South
478 America was characterized by extension during ongoing subduction (Franzese and
479 Spalletti; 2001; Giambiagi et al., 2009; Giacosa et al., 2010; Spikings et al., 2016; González
480 et al., 2017). Its simultaneity with the development of the Viedma basin supports our
481 hypothesis of an extensional aperture of the Viedma basin during the Late Triassic (Fig.
482 10).

483 Folded basement rocks around Paso del Viento that we assign to the NVU on the
484 basis of lithological affinity (Fig. 1) and are overlain in angular unconformity by Upper
485 Jurassic volcanic rocks from El Quemado Complex (153-162 Ma by Pankhurst et al.,
486 2000). Such stratigraphic relations can indirectly constraint the Viedma basin closure,
487 deformation, and metamorphism, between the Late Triassic and the Middle Jurassic (Fig.
488 11). Regionally, this deformational event can be related to the Late Triassic (?)–Early

489 Jurassic Chonide accretionary event (Fig. 11; Thomson and Hervé, 2002; Hervé et al.,
490 2008), which affected the CMC and the Central Patagonian region (Fig. 11; Zaffarana et al.,
491 2014), and was synchronous with the Peninsula orogeny in the Antarctic Peninsula (Fig.
492 11; Vaughan and Livermore, 2005; Hervé et al., 2006). Overall our tectonic reconstruction
493 marks a tectonic mode switching from Late Triassic extension (or transtension?) to Early
494 Jurassic contraction (Fig. 11).

495 In southernmost Patagonia, towards the latest Jurassic new evidences of extension
496 and establishment of a backarc setting are registered in the Rocas Verdes basin (Fig. 11; cf.
497 Calderón et al., 2007a), which located southwest from NVU. This means an oceanward
498 migration of the arc-backarc system after the deformation (closure) of the NVU. Whereas at
499 Central Patagonia also registers, from the Late Triassic to Early Cretaceous, a westward
500 migration of the arc magmatism, in fact, this arc magmatism develops during a period of
501 upper plate extension (Echaurren et al., 2016). This pattern of arc-backarc system younger
502 outboard associated to extension, disrupted by short-lived deformation events, were
503 features identified as typical from retreating orogens (Collins, 2002a, 2002b; Cawood et al.,
504 2009).

505 Even though the dynamics of the Mesozoic orogenic system could be reflecting a
506 retreating-mode, in southernmost Patagonia, its evolution is quite complex and was
507 influenced by the connection with the Antarctic Peninsula (Hervé and Fanning, 2003;
508 Calderón et al., 2007a). The Jurassic arc magmatism emplaced along the Patagonia-
509 Antarctic Peninsula (Hervé et al., 2007; Echaurren et al., 2016; Riley et al., 2016), was
510 synchronous with the Middle-Late Jurassic extensional-transtensional movements that
511 started to separate the Antarctic Peninsula composite terrane (magmatic arc and forearc

512 setting) from Patagonia (König and Jokat, 2006). Indeed, the lithospheric extension finally
513 resulted in the appearance of oceanic floor in the Weddell Sea at ~160-147 Ma (Fig. 11; see
514 discussion in Ghidella et al., 2007) and the partial separation between both blocks,
515 concomitant with the development of the Rocas Verdes basin. Finally, towards Cretaceous
516 times, the rocks of the southernmost Chilean margin were involved in the process of
517 subduction (Hervé and Fanning, 2003; Angiboust et al., 2018) revealing that Antarctic
518 Peninsula was located farther south from Patagonia (Hervé and Fanning, 2003).

519 6. CONCLUSIONS

520 We propose a new lithostratigraphic unit, in the Southern Patagonian Ice Field, and
521 also recognized in the upper parts of the surrounding cordillera (Fig. 1) characterized by
522 very-low grade metamorphic rocks of Late Triassic age. It is the first mention of Triassic
523 metamorphic rocks in Argentina.

524 We discard a volcanic origin to the Nunatak Viedma and we separate these rocks
525 from similar rocks assigned to the Bahía de la Lancha Formation by age and lithology.

526 These Upper Triassic very low- to low-grade metamorphic rocks would integrate a
527 metamorphic inlier that we propose to denominate “Nunatak Viedma Unit”. It is a key
528 element to understand the Patagonia-Antarctic Peninsula geometrical fit and the evolution
529 of the early Mesozoic orogenic system. Given its broad range of detrital ages, it seems that
530 the Viedma basin was probably a extensional retroarc-backarc basin, separated from trench-
531 forearc basins such as the Chonos basin by an active volcanic arc emplaced along Central
532 Patagonia and the Antarctic Peninsula.

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543

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1017 **FIGURE CAPTIONS**

1018 Figure 1. (a) Location and geological map of the study area (modified from Giacosa et al.,
1019 2012b). The yellow star corresponds to the U-Pb detrital zircons sampling site. (b) Zoom-in
1020 on the Nunatak Viedma-Paso del Viento showing with major precision the sample site.

1021 Figure 2. Paleogeographic reconstruction of Pangea at 200 Ma (Triassic-Jurassic boundary)
1022 showing the relative position between Patagonia and Antarctic Peninsula (modified from
1023 the “tight fit model” by Lawver et al., 1998).

1024 Figure 3. Pre-Jurassic representative ages from Patagonia and Antarctic Peninsula. CMC:
1025 Chonos Metamorphic Complex; CPB: Central Patagonian Batholith; CPVC: Curaco
1026 Plutonic-Volcanic Complex. DMC: Darwin Metamorphic Complex; EAMC: Eastern Andes
1027 Metamorphic Complex; MDAC: Madre de Dios Accretionary Complex; NPA: North
1028 Patagonian Andes. NVM: Nunatak Viedma Metamorphites. SOI: South Orkney Islands.
1029 SPA: Southern Patagonian Andes. Patagonian ages are from Saini-Eidukat et al. (2004),
1030 Pankhurst et al. (2006), Zaffarana et al. (2014) and Greco et al. (2017) in the North
1031 Patagonian Massif; Pankhurst et al. (2003, 2006), Moreira et al. (2013) and Permuy Vidal
1032 et al. (2014) in the Deseado Massif; Cingolani and Varela (1976) and Jacobs et al. (1999) at
1033 Malvinas Islands; Hervé and Fanning (2001), Varela et al. (2005), Pankhurst et al. (2006),
1034 Hervé et al. (2016) and Serra-Varela et al. (2016) in the Northern Patagonian Andes and
1035 North Patagonian Chilean coast; Hervé et al. (2003) in the Southern Patagonian Andes;
1036 Sölner et al. (2000), Barbeau et al. (2009b), Calderón et al. (2010), Hervé et al. (2010) and
1037 Castillo et al. (2017) in the Fuegian Andes. Antarctic Peninsula’s ages are from Pankhurst
1038 (1982), Willan et al. (1994), Millar et al. (2002), and Riley et al. (2012).

1039 Figure 4. LA-ICP-MS U-Pb zircon ages for MN-0, MN-1 and MN-10 samples plotted in
1040 Tera and Wasserburg (1972) concordia diagrams. The left column corresponds to the entire
1041 spectrum of ages while the right column shows principally Paleozoic ages.

1042 Figure 5. The left column corresponds to the frequency histograms and relative probability
1043 plots of U-Pb detrital zircon ages (LA-ICP-MS). The inset shows weighted average age
1044 plots of the youngest cluster overlapping at 1σ . Weighted mean age is interpreted as the

1045 maximum depositional age. Pie-chart with the distribution of the zircon detrital ages for
1046 each sample is represented on the right column.

1047 Figure 6. Outcrops of the Nunatak Viedma locality showing lithological and structural
1048 features. A) Very low-grade metapsammites and metapelites with excellent preservation of
1049 bedding cut by a mid angle reverse fault. B) Metasedimentary rocks interbedded with
1050 metabasites with boudins development and calcite veins in the neck of the structure. S₁
1051 dip/dip direction is 60°/340°. C) Metapsammites and metapelites showing upright isoclinal
1052 and tight folds, and high-angle reverse faults.

1053 Figure 7. Photomicrographs of metapsammites and metabasites from NVU. A) typical
1054 composition of the metapsammites. B) S₁ cleavage and associated growth of chlorite +
1055 sericite. C) S₁ cleavage affected by a crenulation cleavage (S₂) developed in brittle
1056 conditions. D) Metabasite with porphyric texture exhibiting plagioclases + clinopyroxene
1057 phenocrysts within fine-grained groundmass composed by mafites, opaques, and
1058 plagioclase.

1059 Figure 8. Paleozoic-lower Mesozoic relative probability distribution from Chonos
1060 Metamorphic Complex (CE9603, FO9606 and CE9625 samples from Hervé and Fanning,
1061 2001) and Viedma Nunatak Unit (MN0, MN10 and MN1 samples from this study). NVU
1062 shows prominent lower Paleozoic peaks, which are not represented in the CMC, while the
1063 Permian-Triassic peaks are represented in both units. Violet and sky-blue lines depict the
1064 MDA from CMC and NVU, respectively.

1065 Figure 9. Samples from the Chonos Metamorphic Complex and the Nunatak Viedma Unit
1066 are plotted as curves of cumulative proportion based on differences between the
1067 crystallization and depositional ages (CA – DA) of the detrital zircons. This diagram
1068 enabling prediction of the tectonic setting of metasedimentary packages of unknown origin
1069 (see procedure in Cawood et al., 2012). Both metamorphic units present typical curves of
1070 convergent setting while the Chonos Metamorphic Complex presents a typical curve for
1071 forearc or trench basins, the Nunatak Viedma Unit presents a curve akin to backarc basins.
1072 Brown areas represent overlapping tectonic fields.

1073 Figure 10. Interpretation of the Late Triassic tectonic setting of the southwestern margin of
1074 Pangea and the paleogeographic implications on the Patagonia-Antarctic Peninsula
1075 connection. The A-A' schematic profile exhibits the tectonic setting south of the 47°S, and
1076 the developed of the Nunatak Viedma basin in a retroarc position respect to the magmatic
1077 arc emplaced along Antarctic Peninsula. Subsequent closure of the basin occurred during the
1078 Chonide deformation event. The B-B' schematic profile (based from Rapela et al., 1992;
1079 Zaffarana et al., 2014, 2017) show the tectonic setting north of the 47°S. The red line in the
1080 paleogeographic reconstruction depicts the position of the magmatic arc (topographic
1081 barrier) emplaced along Antarctic Peninsula arc with continuation in Patagonia (Central
1082 Patagonian Batholith and farther north). EAMC: Eastern Andes Metamorphic Complex.
1083 GSF: Greywacke-Shale Formation.

1084 Figure 11. Correlation between units from Patagonia and Antarctic Peninsula and the
1085 tectonic events happened during the early Mesozoic in the southwestern margin of Pangea.
1086 Besides, we exhibit representative geochemical features of the volcanic and plutonic units.
1087 Based on Dalziel (1982), Trouw et al. (1997), Thomson and Hervé (2002), König y Jokat
1088 (2006), Calderón et al. (2007, 2013), Hervé et al. (2007), Riley et al. (2010), Vaughan et al.
1089 (2012), Zaffarana et al. (2014), Echaurren et al. (2016), Ghiglione et al. (2016) and this
1090 study. V1, V2, and V3 represent episodes of Jurassic volcanism defined by Pankhurst et al.
1091 (2000).

1092 **APPENDIX A. SUPPLEMENTARY DATA.**

1093 Table A. LA-ICPMS U-Pb geochronological analyses in detrital zircons of the samples
1094 MN-0, MN-1, and MN-10.

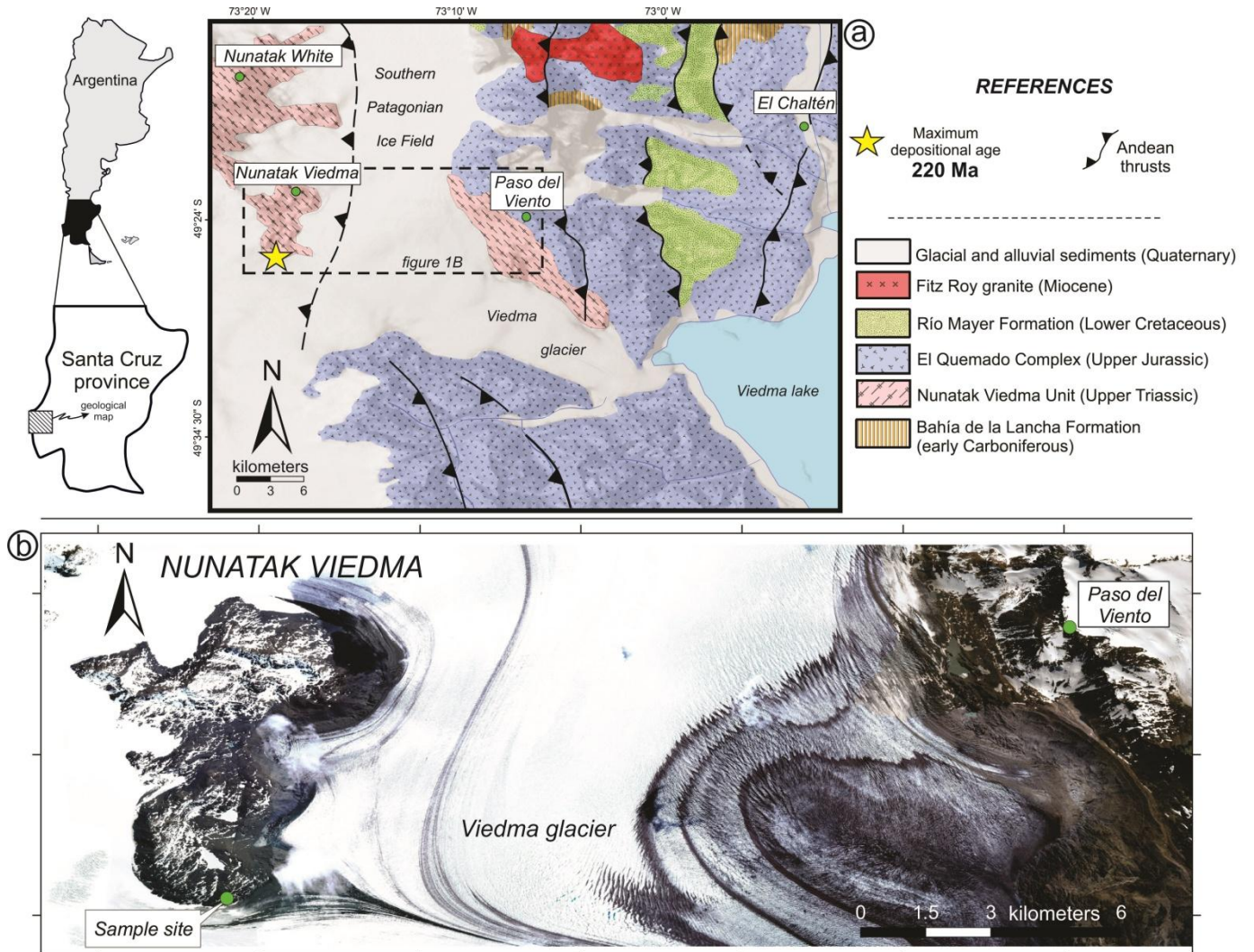


FIGURE 1

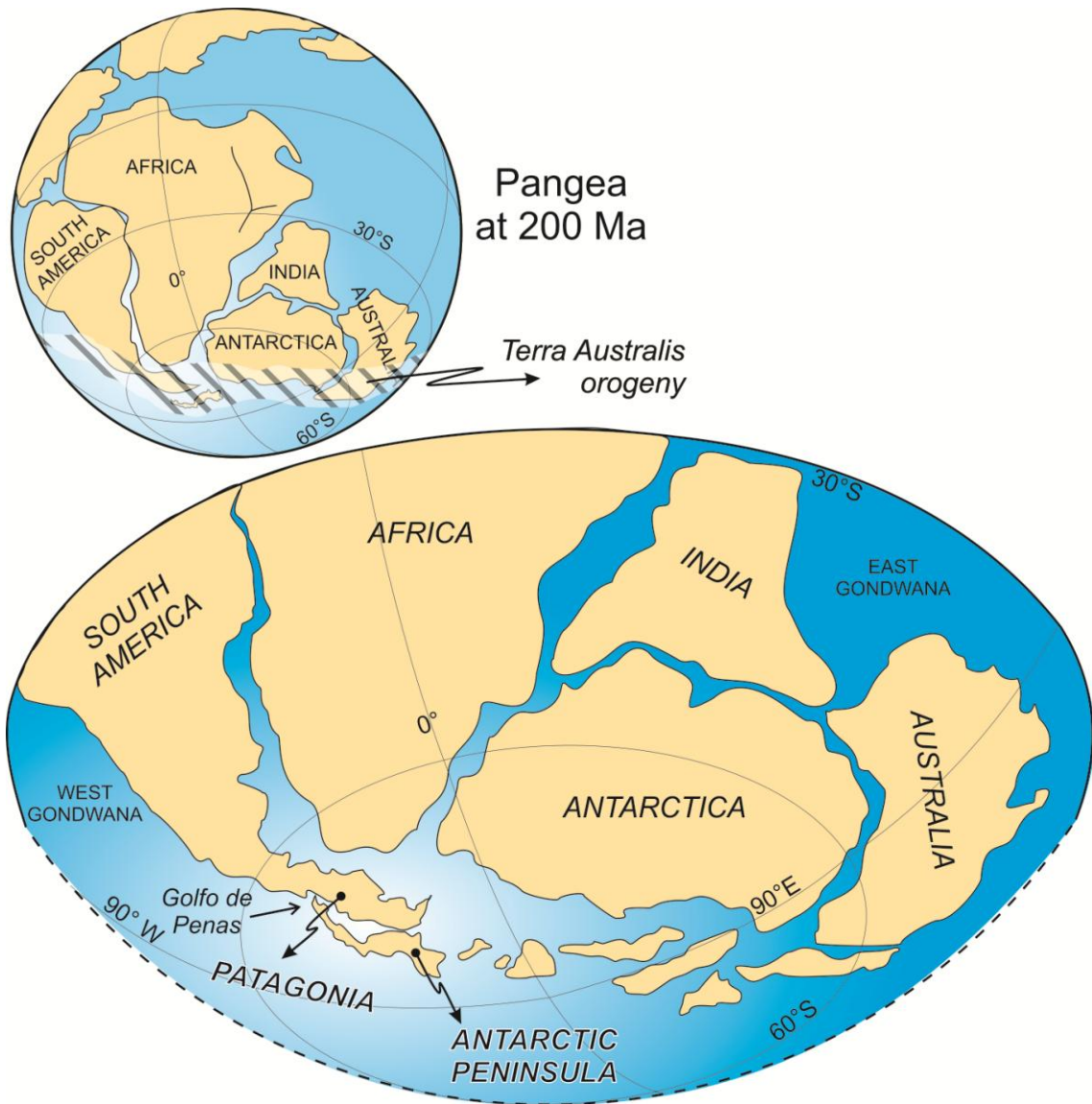


FIGURE 2

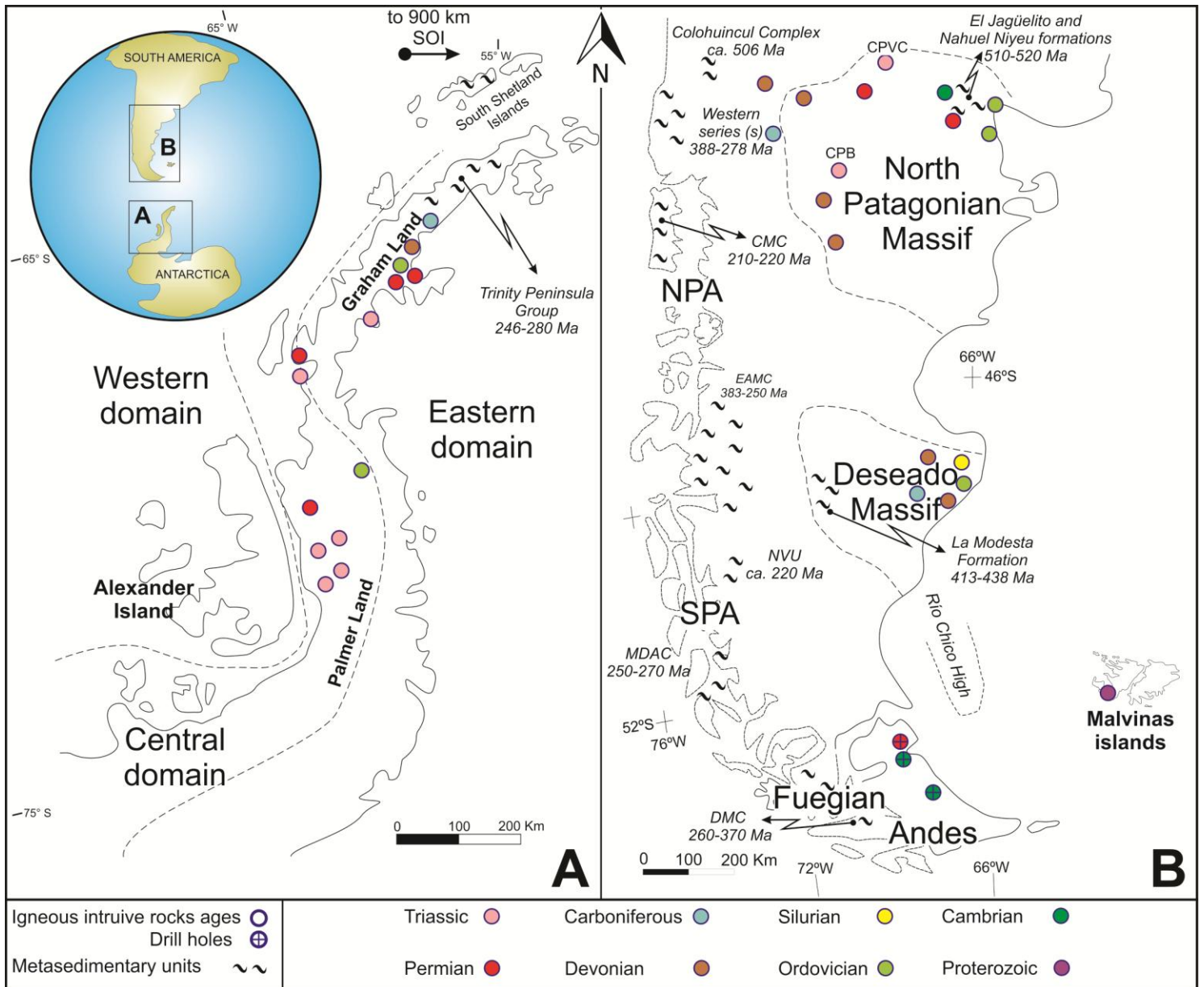


FIGURE 3

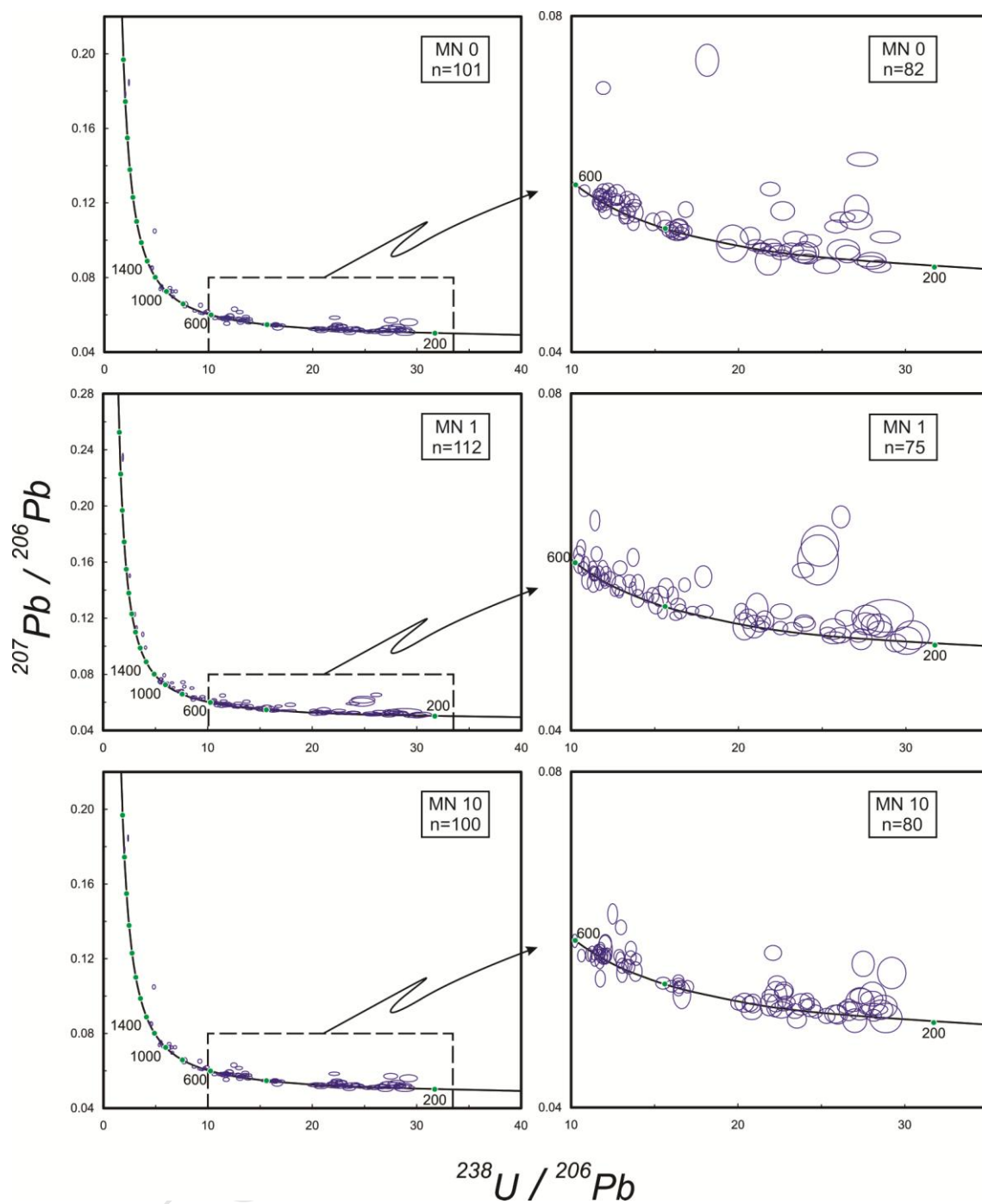


FIGURE 4

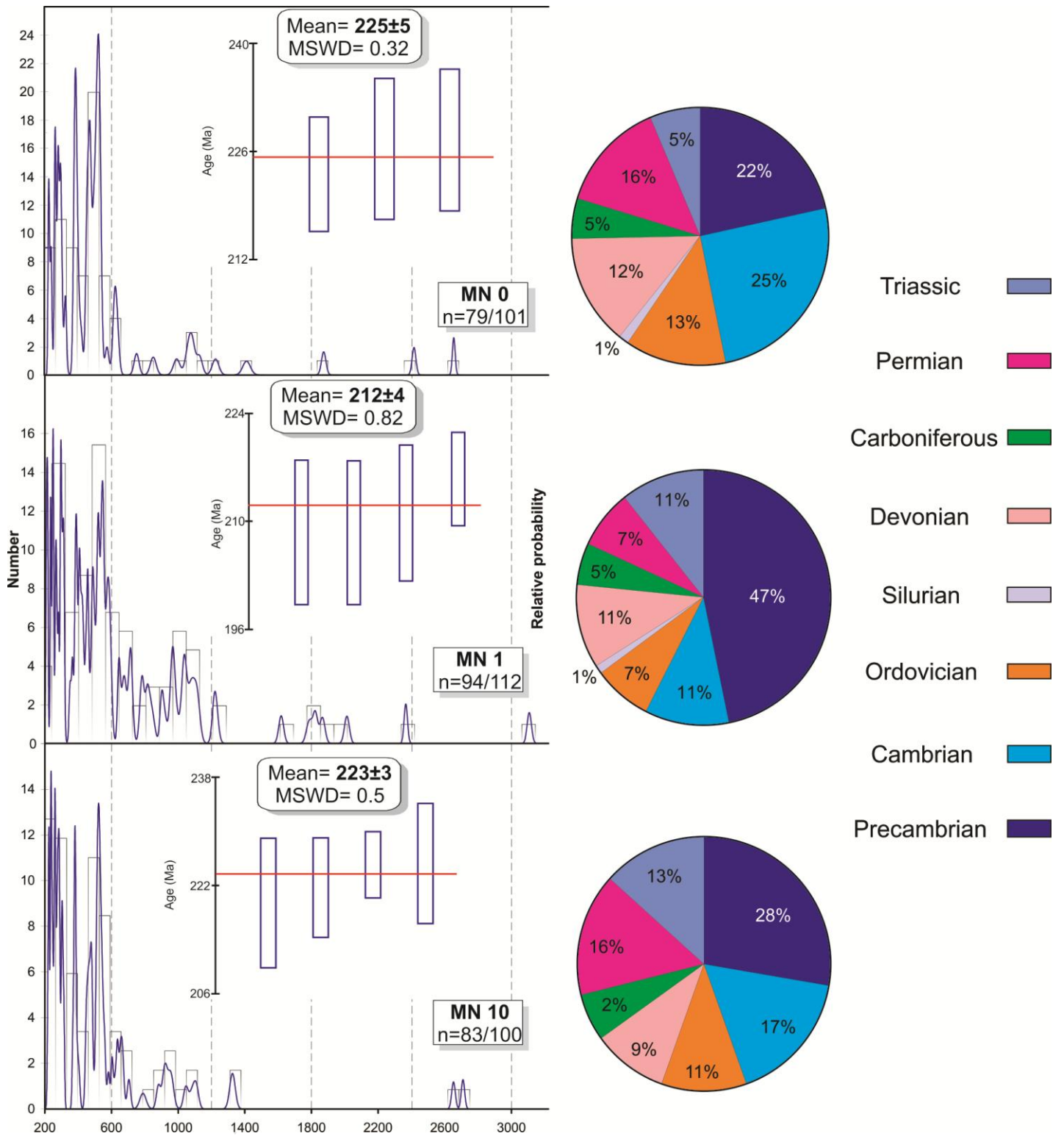


FIGURE 5

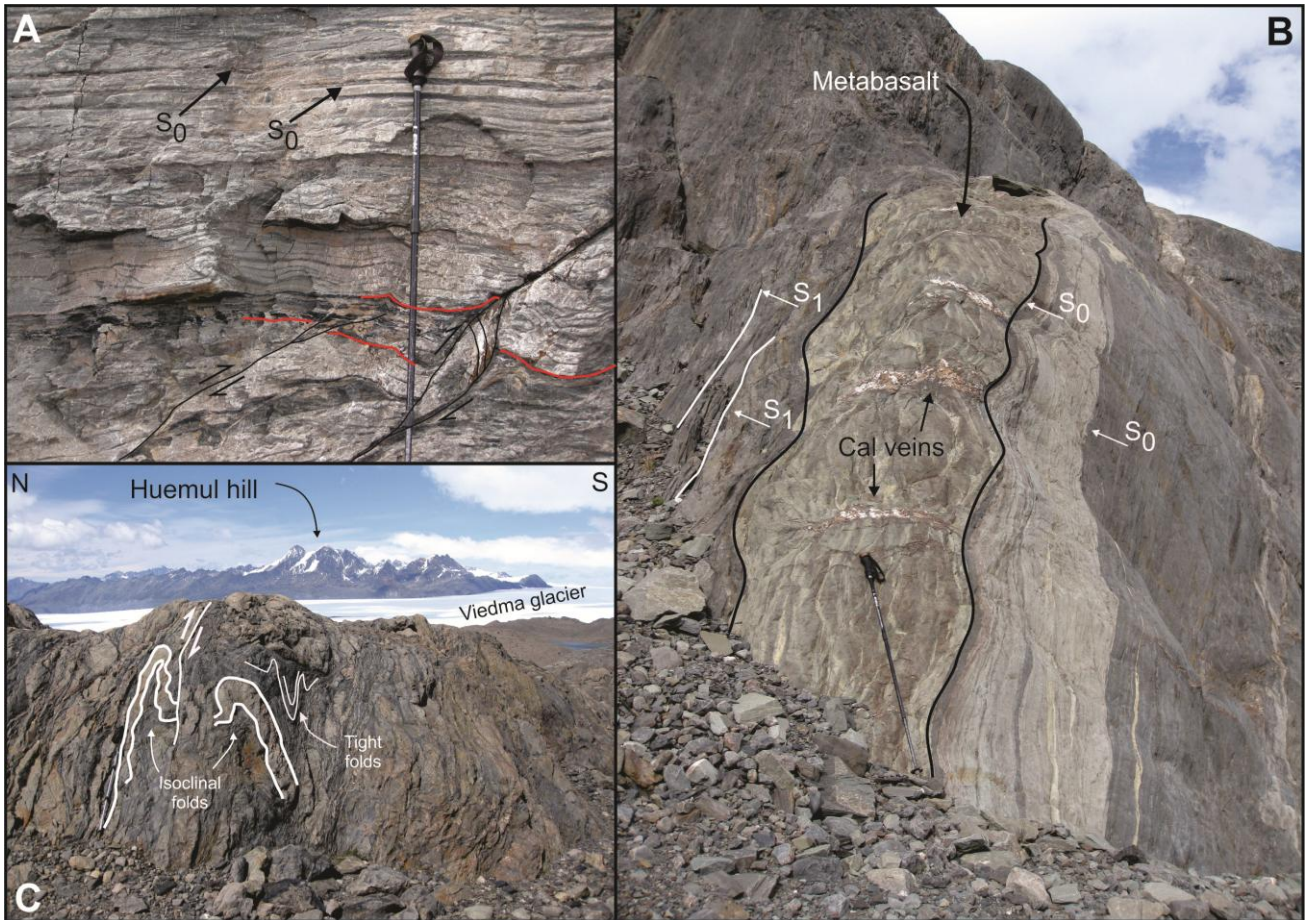


FIGURE 6

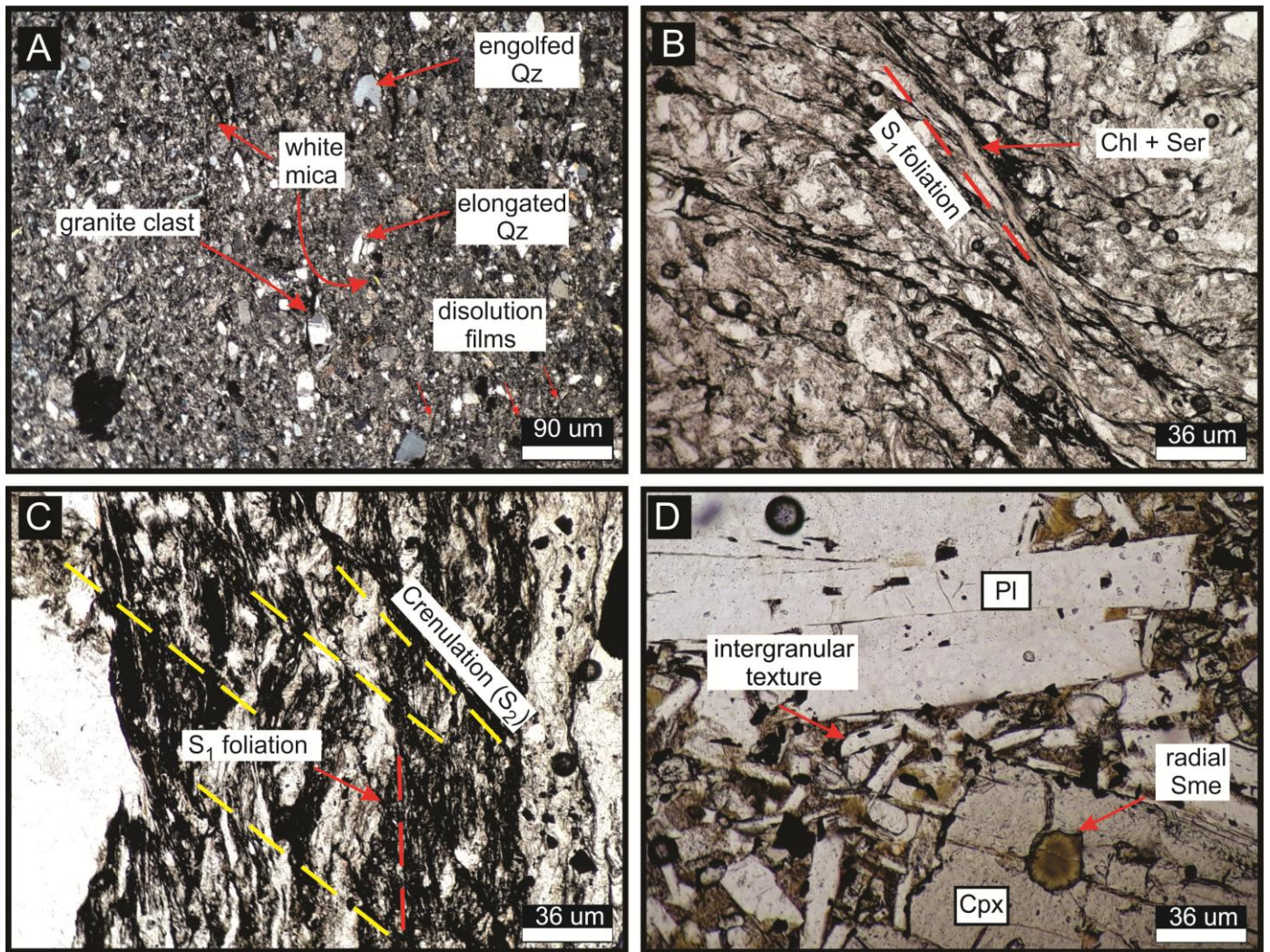


FIGURE 7

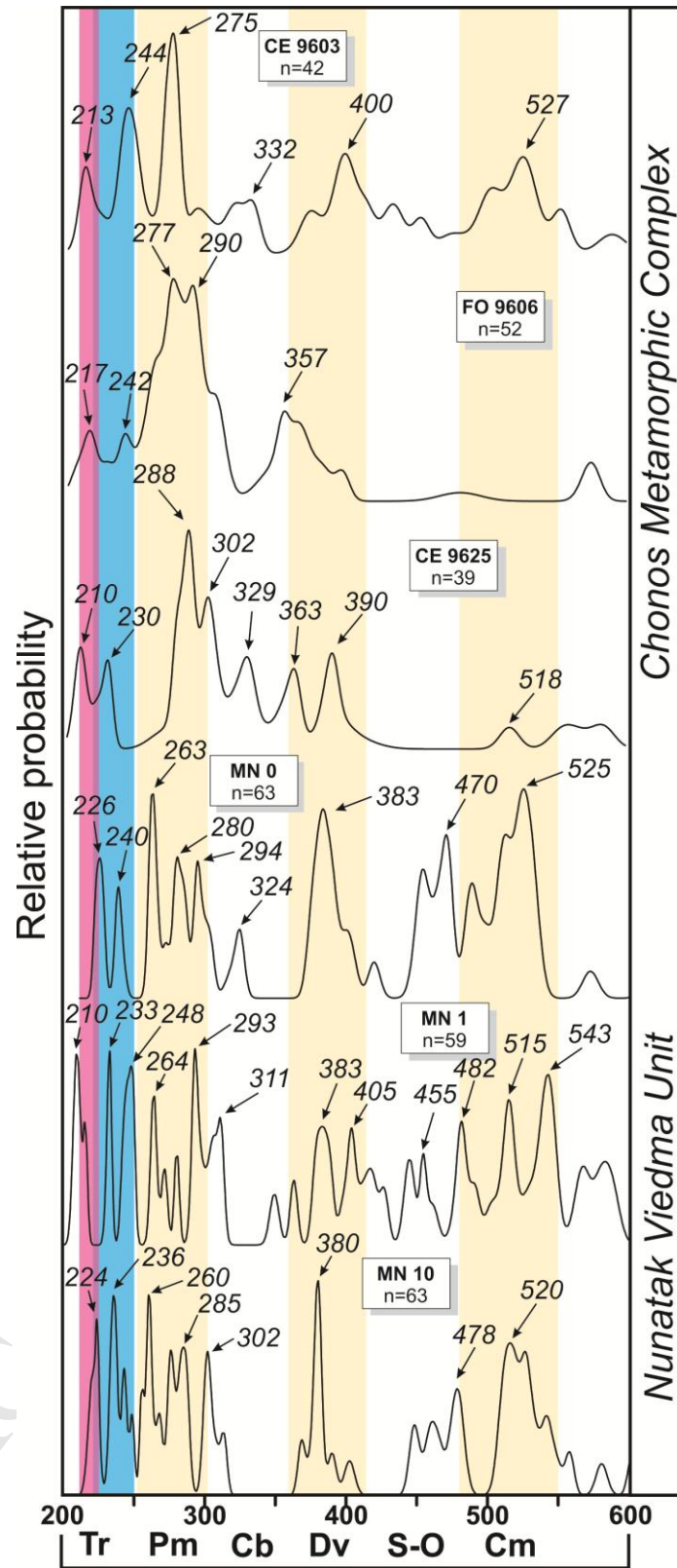


FIGURE 8

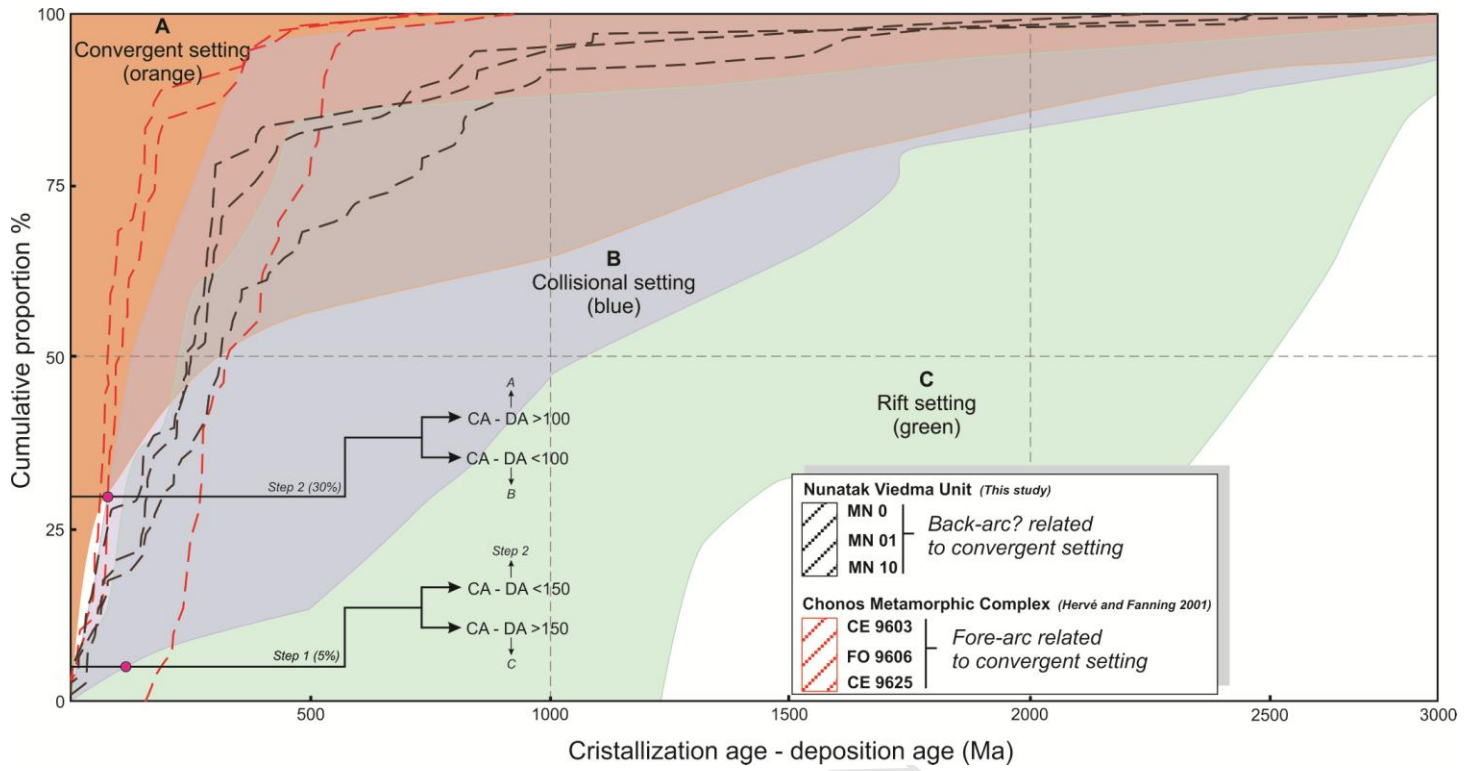


FIGURE 9

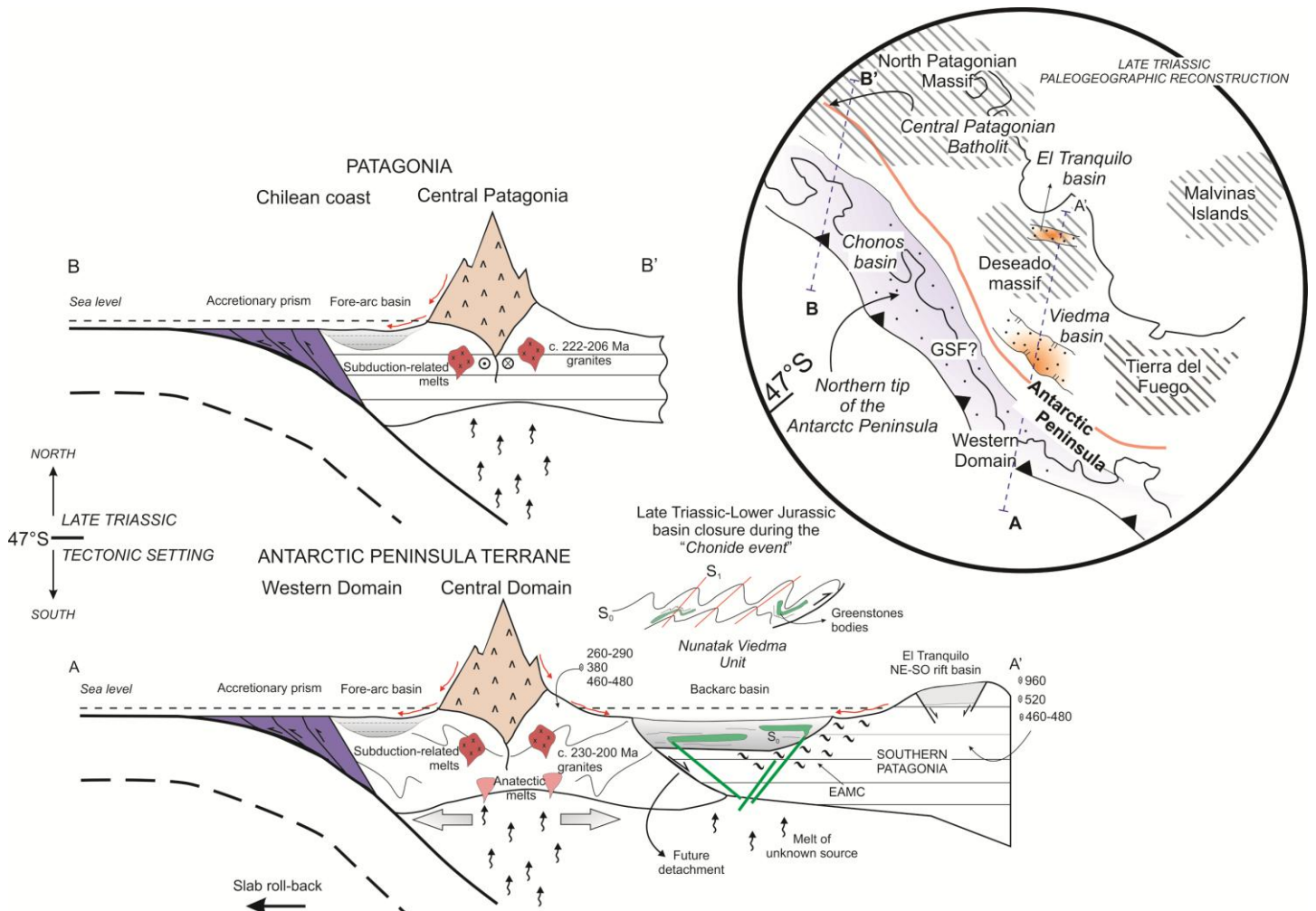


FIGURE 10

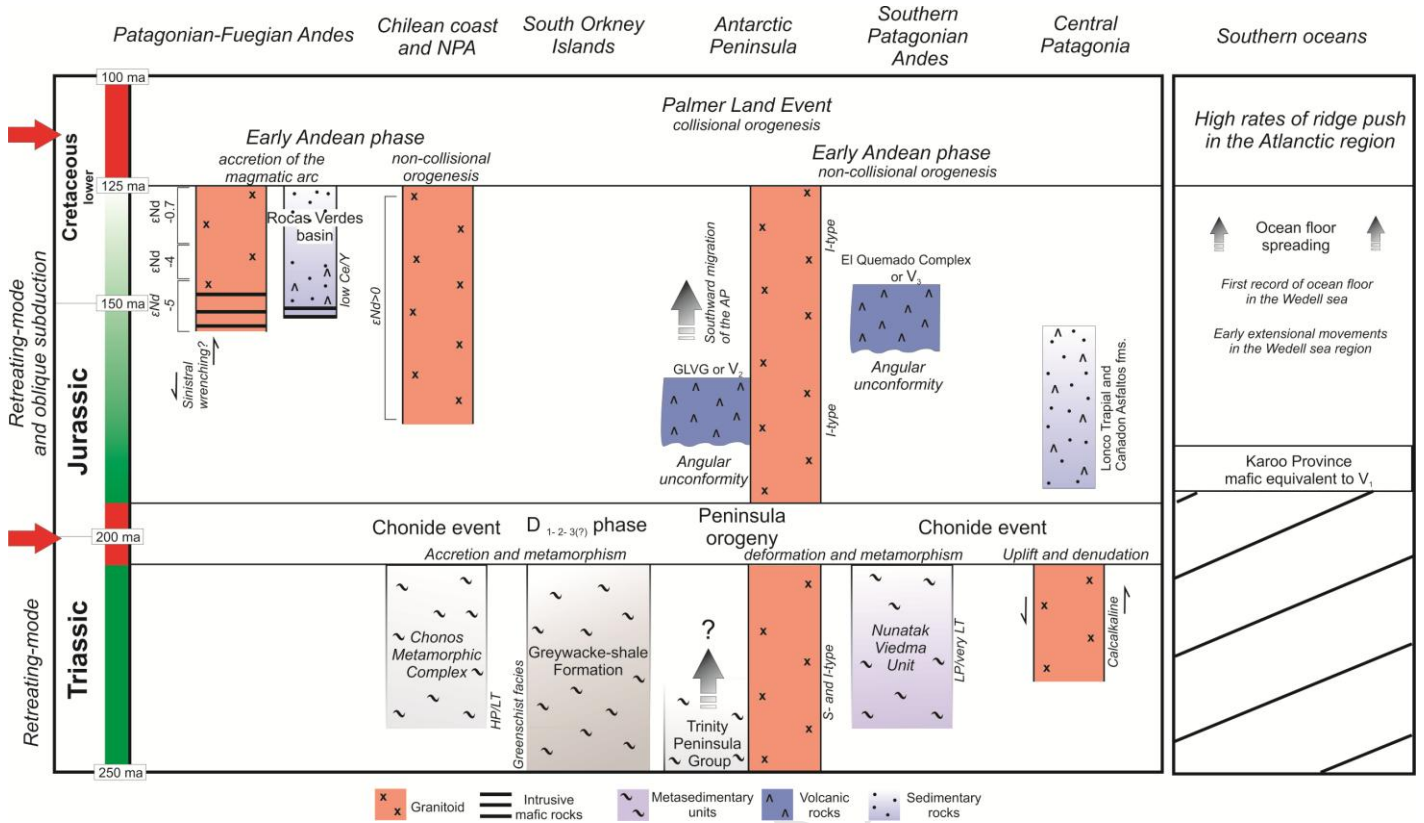


FIGURE 11

RESEARCH HIGHLIGHTS

- We recognized very-low grade metamorphic rocks on the Southern Patagonian Icefield.
- Late Triassic maximum depositional ages were constrained by zircon detrital ages.
- We propose to define the Nunatak Viedma Unit.
- A backarc setting is interpreted.
- Detrital zircon ages support Antarctic Peninsula-Patagonia Triassic connection.